belong to the surviving ones with the exception of large dasycladaceae or genus *Cymopolia* sp. (4) Biostratigraphic determination of Lower Danian beds is confirmed by the species *Globorotalia compressa* and *Protelphidium* sp. in the Sopada section. By this the carbonate complex of beds can be compared for the first time with deeper marine beds on the basis of planktonic zonation. (5) By further biotic and abiotic studies the shallow marine carbonates of the Adriatic platform will have to be more successfully documented and linked with the global sequence of events at the KT boundary.

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THE END-PERMIAN MASS EXTINCTION: A COMPLEX, MULTICAUSAL EXTINCTION. D. H. Erwin, Department of Paleobiology, NHB-121, National Museum of Natural History, Washington DC 20560, USA.

The end-Permian mass extinction was the most extensive in the history of life and remains one of the most complex. Understanding its causes is particularly important because it anchors the putative 26-m.y. pattern of periodic extinction. However, there is no good evidence for an impact and this extinction appears to be more complex than others, involving at least three phases [1,2]. The first began with the onset of a marine regression during the Late Permian and resulting elimination of most marine basins, reduction in habitat area, and increased climatic instability; the first pulse of tetrapod extinctions occurred in South Africa at this time. The second phase involved increased regression in many areas (although apparently not in South China) and heightened climatic instability and environmental degradation. Release of gas hydrates, oxidation of marine carbon, and the eruption of the Siberian flood basalts occurred during this phase. The final phase of the extinction, episode began with the earliest Triassic marine regression and destruction of nearshore continental habitats. Some evidence suggests oceanic anoxia may have developed during the final phase of the extinction, although it appears to have been insufficient to be the sole cause of the extinction.

Some 50% of marine families and perhaps over 90% of marine species were eliminated during the last few million years of the Permian [1,3]. Preservational problems and geochemical evidence suggest the extinction may have persisted into the earliest Triassic. Extinction patterns were highly selective: conodonts, nonfusulinid foraminifera, most bivalves, nautiloid cephalopods, and bellerophontid gastropods were largely unaffected, while blastoids, camerate crinoids, and tabulate corals had been declining well before the end of the period and disappeared entirely by the latest Permian. The last trilobites, stenolaemate bryozoans, rugose corals, and crinoids persisted at low diversity into the final, Changxingian stage. The extinction primarily affected the fusulinid foraminifera, several groups of articulate brachiopods, and ammonoid cephalopods; cidaroid echinoids, other gastropods, sponges, and some brachiopod groups declined in the latest Permian, but rebounded fairly quickly during the Early Triassic.

Late Permian faunas are highly provincial, however, and the biogeographic component of the extinction is only now being assessed. The best data for rapid extinction come from South China [4–6], an isolated tectonic block during the late Permian. It remains unclear to what extent the data from South China are representative of global diversity patterns. Biogeographic, taphonomic, and other problems continue to plague analysis of extinction data, obscuring important information on the rate and pattern of extinction. Recently developed conodont biostratigraphic correlations hold the promise of answering such questions [7]. It does seem fairly clear that sessile, epifaunal filter-feeders, and nearshore taxa suffered higher extinction rates than other taxa.

On land, 21 of 27 (75%) tetrapod families disappeared [8]. In the Karoo Basin of South Africa two extinction peaks have been described from the Late Permian [9] although sampling problems raise important questions. Of the 27 orders of insect recorded from the Permian, eight disappeared and an additional seven suffered considerable reductions; four of the latter recovered during the Triassic. This was the most extensive ordinal extinction in insect history [10]. Plants show little evidence for mass extinction, although both pollen and fungal spores show significant changes at the Permo-Triassic boundary [1,11]. These changes may be associated with the destruction of nearshore communities during the rapid Early Triassic transgression.

Determining the cause of the extinction is complicated by the multitude of Late Permian geological changes. Although the supercontinent of Pangea had formed tens of millions of years earlier, a widespread regression during the terminal Permian exposed most of Pangea [12] leading to widespread climatic instability [13]. The rapid eruption of the Siberian flood basalts [14], significant shifts in C, O, S, and Sr isotopes [1,12,15,16], and increasing evidence for some marine anoxia [17] complicate the picture. The $\delta^{13}C$ evidence has received considerable attention, but appears to be equally consistent with erosion and oxidation of marine carbonates [12,15], formation of anoxic bottom waters [17–20], and release of methane gas hydrates during regression [1].

Recently proposed causes for the end-Permian extinction include volcanism-induced cooling, extraterrestrial impact, and global anoxia (reviewed in [1]). Although no Ir anomaly has been identified at the P/Tr boundary [21], microspherules, apparently of volcanic origin, are widespread [e.g., 4]. However, despite the apparent force of the eruption of the Siberian traps, the eruption appears to have begun after the extinction. Additionally, the eruption appears to have been insufficient to trigger an extinction this large. Anoxia hypotheses have a related problem: calculations suggest that the observed $\delta^{13}C$ shift may have been insufficient to reduce atmospheric O levels sufficiently to induce extinction [1,2].

The most plausible current explanation of the end-Permian mass extinction appears to involve multiple elements in a complex extinction, as outlined above. Testing this scenario, and the others proposed, requires high-resolution, global analyses of survival and extinction with particular attention to various biases (taphonomic, taxonomic, etc.), which may bias the data. Recent biostratigraphic advances [7] now make this possible and such studies are currently underway in the U.S. and China.

The end-Permian mass extinction differs sharply from the end-Cretaceous in length, the pattern of extinction, and in the complexity of associated geological events. Additionally, there is little evidence for extraterrestrial impact at the close of the Permian and C, O, S, and Sr isotopic patterns are significantly different [22]. These sharp differences pose important problems for attempts to link these two extinctions to a common cause.

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NEW EVIDENCE FOR PRIMARY FRACTIONATION OF RUTHENIUM AND IRIDIUM IN THE CHICXULUB EJECTA CLOUD.

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Introduction: Platinum-group element (PGE; Ru, Ir, Pt, Pd, Ir, plus Au, included in the term PGE for the sake of brevity) analysis of the KT boundary fireball layer (terminology of [1]) has revealed that the mean Ru/Ir ratio of the many marine sites studied (1.67 ± 0.38) is close to the chondrite ratio (1.48; [2]), whereas the value for nonmarine sites (0.76 ± 0.26) is not [3,4]. A positive correlation between the Ru/Ir ratio of globally distributed KT sites and 65-Ma paleodistance from the Chicxulub structure was also observed [4]. This trend suggested that temperature-dependant, primary fractionation of PGE occurred in the ejecta cloud during condensation of vaporized projectile material. However, this previous work could not negate the dependence of the paleodistance-Ru/Ir ratio correlation on environment of deposition because all the marine sites studied were far from Chicxulub (Europe and New Zealand) and all the nonmarine sites were in North America. This work presents additional Ru and Ir analysis of the fireball layer from marine KT sites proximal to the impact structure at Chicxulub, Yucutan, other North American nonmarine KT sites, and DSDP cores. With this new data we assess the dependence of the previously observed trend on depositional environment and suggest a simple mechanism for primary fractionation of PGE, prior to their deposition in the fireball layer.

Methods and Results: Ruthenium and Ir in KT boundary samples were measured by isotope dilution inductively coupled mass spectrometry (ICP-MS), and a complete description of the digestion and analytical procedures have been published elsewhere [3,4]. Precision for the technique as percent standard deviation for multiplicate analysis of USGS diabase standard W-1 are 6% (Ru) and 0.7% (Ir).

The abundances of Ru and Ir for all KT sites is given in Table 1. The Ru/Ir ratio and nature of the fireball layer samples analyzed from the following sites have been previously published [4]; Petriccio, Italy; Knappengraben and Elendgraben, Austria; Stevns Klint, Denmark; Agost, Spain; Woodside Creek, New Zealand; Raton Basin, Colorado; Red Deer Valley, Alberta; Morgan Creek, Saskatchewan and Lance Creek, Wyoming. New data is presented here for DSDP 577, DSDP 596, Brazos River, Texas, and Beloc, Haiti.

TABLE 1. Paleodistance to Chicxulub and Ru/Ir ratio for the fireball layer at the KT boundary sites studied.

Site	65 Ma Paleodistance to Chicxulub	Ru/Ir	Ru (ppb)	Ir (ppb)
Raton Basin				
Berwind Canyon	52	0.89	50.9	56.9
Starkville South	52	0.97	53.0	54.5
Clear Creek North	52	0.59	8.4	14.2
Lance Creek	2635	1.1	2.7	2.4
Morgan Creek	3440	0.43	7.8	18.0
Frenchman River	3532	0.49	1.4	2.9
Hell Creek	3299	0.60	0.83	1.4
Red Deer Valley	3711	0.57	2.4	4.2
Stevns Klint	7363	1.6	69.4	42.9
Agost	6838	1.9	44.1	23.9
Petriccio	8318	1.2	4.6	3.7
Knappengraben	7730	2.0	4.9	2.5
Elendgraben	7642	2.2	18.7	8.4
Woodside Creek	12019	1.4	36.8	25.9
Brazos River	1338	0.87	0.76	0.85
Beloc, Haiti	688	0.72	4.1	5.7
DSDP 577	9621	1.7	0.47	0.27
DSDP 596	9378	0.8	8.0	9.7
Chancet Rocks*	12019	2.4	11	4.6
Needles Point*	12019	2.9	19	6.4

nd = no data; All analyses are on a carbonate-free basis. Stevns Klint sample is of the lower layer III of [16]; spheroid-rich red clay, 0–1 cm above the top of the Maastrichtian. DSDP 577 sample is from Leg 86, core 12, section 5, 115–116 cm. DSDP 596 value is a mean of eight 1-cm continuous samples over the boundary interval; leg 91, core 32, section 4, 47–55 cm. Paleoreconstruction is courtesy of the PLATES PROJECT, Institute for Geophysics, The University of Texas at Austin.

*Data from [11]. Petriccio sample is of the middle 0.5 cm of the boundary clay (red layer).